**A model of the 4,000-year paleohydrology ($\delta^{18}\text{O}$) record
from Lake Salpetén, Guatemala**

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Abstract

A simple mass-balance model provides insights into the influence of catchment vegetation changes and climate variability on the hydrologic and stable oxygen isotope ($\delta^{18}\text{O}$) evolution of Lake Salpetén, in the Maya Lowlands of northern Guatemala. Model simulations for the last 4,000 years incorporate pollen-inferred changes in vegetation cover and account for 75% of the variance observed in the biogenic carbonate $\delta^{18}\text{O}$ record from a long lake sediment core. Vegetation-driven hydrologic changes, however, failed to capture the full range of late Holocene sediment core $\delta^{18}\text{O}$ variability. The model requires incorporation of additional shifts in catchment vegetation cover, inclusion of regional precipitation changes, or likely both, to explain the fluctuations observed in the lake core oxygen isotope record. Climatic interpretation of the model results suggests there was relatively greater moisture availability between about 2,400 and 1,800 years ago, but increased $\delta^{18}\text{O}$ values centered at ~3,300, 2900, 500, and 200 calendar years before present (cal yr BP) indicate abrupt precipitation decreases. There is evidence for protracted aridity between 1,500 and 800 cal yr BP.

1. Introduction

Over the last several decades, interest in late Holocene paleoclimate of the Maya Lowlands has grown. Earth scientists have exploited a number of natural archives of past climate information, including speleothems (cave stalagmites) (Medina-Elizalde et al., 2010, Medina-Elizalde and Rohling, 2012, Kennett et al., 2012 and Webster et al., 2007), marine sediments (Haug et al., 2003) and lake deposits (Hodell et al., 1995, 2005a, 2005b, 2007, 2012, Curtis et al., 1996, Douglas et al., 2014 and Rosenmeier et al., 2002a, 2002b) to infer long-term variability in regional rainfall amount, and thereby provide a climate context for ancient Maya cultural development and decline. Pollen in lake sediment cores from the region is largely uninformative about late Holocene climate change because ancient inhabitants cleared much of the forest for agriculture and construction (Vaughan et al., 1985, Leyden, 2002, Wahl et al., 2006). Thus, paleolimnological studies of late Holocene climate in the Maya Lowlands have relied largely on stable oxygen isotope ($\delta^{18}\text{O}$) values in carbonate shells of snails (molluscs) and ostracods (crustaceans), which were thought to be immune from the influence of human activities. The rationale for this approach has been spelled out in detail elsewhere (Talbot, 1990, Chivas et al., 1993, Curtis and Hodell, 1993, Holmes 1996, Brenner et al., 2003, and Douglas et al., this issue), but is summarized briefly here.

1.1. Rationale for isotope-based paleoclimate studies in the Maya Lowlands

Oxygen isotope values in carbonate shells from lakes are largely controlled by the $\delta^{18}\text{O}$ of the lake water, the temperature at which the shell carbonate (calcite or aragonite) precipitation occurred, and taxon-specific fractionation. Most isotope-based paleolimnological studies of past climate utilize shell material from adult specimens of a single taxon to eliminate potentially confounding differences in fractionation between species, or between developmental stages of a single species. In cases for which a single taxon is not encountered throughout the sediment record, multiple taxa can be used, but it is helpful to assess inter-species differences in isotopic equilibrium, which can sometimes be achieved by comparing $\delta^{18}\text{O}$ values in shells of the two taxa at depths where they co-occur in the sediment profile.

Temperature is assumed to have had minimal influence on shell carbonate $\delta^{18}\text{O}$ in Holocene lake sediment cores from the Maya Lowlands, in part because infaunal (burrowing) and epifaunal (bottom-dwelling) gastropods and ostracods experience near-constant temperatures at the sediment-water interface year-round, particularly in deeper lakes (Perez et al., 2012). Intra-annual surface-water temperature variations in the region are also relatively small (typically $<5\text{ }^{\circ}\text{C}$), suggesting that shallow-water taxa, likewise, experience a limited range of temperatures throughout their lives.

Perhaps more importantly, emerging evidence indicates that inter-annual (year-to-year) temperature variability in the Maya Lowlands was also limited during the late Holocene. Specifically, new dual-isotope measurements on paired shell aragonite and gypsum samples from Lake Chichancanab, Mexico, suggest that mean temperature ($26\text{ }^{\circ}\text{C}$) and

temperature ranges (23 to 29.5 °C) in the northern Yucatán Peninsula *ca.* 1,200 years ago were similar to values in the region today (Hodell et al., 2012). Thus, changing lake water $\delta^{18}\text{O}$ was probably the principal control on shell carbonate $\delta^{18}\text{O}$ in regional lakes during the late Holocene.

One determinant of lake water $\delta^{18}\text{O}$ is the isotopic signature of hydrologic input water, i.e. rainfall plus surface and sub-surface runoff. The $\delta^{18}\text{O}$ of rainfall can vary as a function of temperature, relative humidity, and changing source water. Assuming that input water $\delta^{18}\text{O}$ values were relatively constant throughout the late Holocene, shifts in the hydrologic budget of the lake, i.e. relative changes in inputs to outputs (evaporation plus surface and sub-surface outflow), were the primary determinant of the oxygen isotope ratio of lake water. Most lakes in the tropical karst terrain of the Maya Lowlands lack surface outflows. In such “closed-basin” tropical water bodies, it is generally assumed that the major control on the lake hydrologic budget is the relation between evaporation (E) and precipitation (P) (Lister et al., 1991). Whereas this assumption is appropriate for lakes in many tropical areas, it may not be fully justified in the Maya Lowlands, where ancient, human-mediated deforestation probably altered watershed hydrology, thereby invalidating the assumption that stratigraphic shifts in the $\delta^{18}\text{O}$ of shells in lake sediment cores reflect solely E/P fluctuations.

Variations in the oxygen isotopic composition ($\delta^{18}\text{O}$) of biogenic carbonate in sediment cores from Lake Salpetén, Guatemala, for example, were used to infer past environmental changes in the southern Maya Lowlands of the Yucatán Peninsula (Rosenmeier et al.,

2002a, 2002b). Results indicated there had been pronounced changes in watershed hydrologic balance over the last ~4,000 years. Strictly climatic interpretation of the $\delta^{18}\text{O}$ record suggested higher precipitation during the early period of Maya settlement expansion, between 2,400 and 1,800 cal yr BP, corresponding roughly to the Late Preclassic (Rosenmeier et al., 2002a, 2002b). Alternatively, this ~600-yr period of low $\delta^{18}\text{O}$ values may have been a consequence of increased surface runoff and groundwater inflow to the lake, related to watershed deforestation by the ancient Maya. Population in the southern Maya Lowlands began to decline at the end of the Late Classic period, *ca.* AD 900, as $\delta^{18}\text{O}$ values in sedimented biogenic carbonates increased as a consequence of reduced hydrologic input to the lake, associated with decreased precipitation or forest recovery.

1.2 Study objectives

Isotope mass balance approaches have been used to evaluate past hydrologic changes in large and small lakes (Hostetler and Benson, 1994, Ricketts and Johnson, 1996, Cross et al., 2001, Rowe and Dunbar, 2004, Jones et al., 2007, and Steinman et al., 2010, 2013a, 2013b). We applied this method to Lake Salpetén to assess quantitatively the impact of catchment vegetation changes and climate on lake level and lakewater $\delta^{18}\text{O}$. Model simulations of lake hydrologic inputs and outputs (and associated $\delta^{18}\text{O}$ values) were constrained by modern meteorological and catchment data, and palynologically inferred changes in forest cover. These simulations tested the plausibility that systematic changes in late Holocene precipitation controlled the hydrologic and isotopic evolution of the lake.

Model results were compared with $\delta^{18}\text{O}$ variations in ostracod and gastropod shells from sediment cores taken in Lake Salpetén and other water bodies on the Yucatán Peninsula.

1.3. Study site

The karst limestone landscape of Petén, northern Guatemala (Fig. 1a), is generally characterized by well-drained forest soils and tropical semi-deciduous and evergreen vegetation (Lundell, 1937). Regional terrain varies in elevation between approximately 100 and 300 m above sea level and the water table lies deep below the land surface. Surface waters, however, are perched, resulting in numerous lakes and seasonally inundated topographic depressions in the landscape (Deevey et al., 1980). The Petén Lake District contains a number of terminal basins distributed along a series of east-west aligned faults at $\sim 17^\circ\text{N}$ latitude (Fig. 1a). Principal water bodies of the lake chain extend approximately 100 km from westernmost Lake Perdida, eastward to the twin basins of Lakes Yaxhá and Sacnab (Brenner et al., 2002). Lake Salpetén, located near the center of the lake chain, is a small water body (2.55 km^2) with a catchment area of 6.36 km^2 (Fig. 1b) and no surface inflows or outflows. The lake lies $\sim 104\text{ m}$ above sea level and has a maximum depth of $\sim 32\text{ m}$ (Anselmetti et al., 2007).

Figure 1: Rosenmeier et al. ↑

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Fig. 1.

(a) Map of Petén, northern Guatemala, showing lake study sites (labeled) and meteorological stations discussed in the text: (1) El Paso, (2) Flores, (3) Libertad, (4) Poptún, (5) Porvenir, (6) San Andrés, and (7) San Luis. (b) Bathymetry and topography of Lake Salpetén and the surrounding watershed. Topographic and bathymetric contour lines appear at intervals of 20 and 3 m, respectively.

Rainfall in Petén varies from ~1,500 to 2,100 mm yr⁻¹ (Table 1). The rainy season occurs between late May and October and is associated with northward migration of the inter-tropical convergence zone (ITCZ) and the North Atlantic high-pressure system (Hastenrath, 1984). During this period, tropical low-pressure systems are carried westward by the trade winds across the Atlantic into the Caribbean, at times bringing heavy rainfall, lightning, and strong winds to the Yucatán Peninsula (Wilson, 1980). Drier conditions develop in November and December and persist through early May, as the ITCZ and North Atlantic subtropical high (Azores-Bermuda high) move southward and atmospheric subsidence becomes predominant. Mean monthly air temperatures in Petén vary between ~20 and 29°C (Table 1), but diurnal temperatures display a greater range of variability.

2. Materials and methods

2.1. Model description

The hydrologic balance of a lake (∂V_L) is controlled by the transfer of water to and from the catchment, according to the equation:

$$\partial V_L = \Sigma I + P - \Sigma O - E \quad (1)$$

where ΣI and ΣO are the total surface and sub-surface inflows (I) to and outflows (O) from the lake, P is direct precipitation on the lake, and E is the evaporative loss from the lake (Dinçer, 1968, and Gat, 1981). A similar expression can be written for the oxygen isotopic composition of lake water (δ_L):

$$\partial V_L \delta_L = \Sigma I \delta_I + P \delta_p - \Sigma O \delta_O - E \delta_E \quad (2)$$

where δ is the isotopic composition of the various inputs and outputs. In lakes with negligible surface outflow, the hydrologic balance and lake water isotopic composition are determined by the difference between evaporation and precipitation over the lake, the water balance of the surrounding catchment (ΣI), and losses through sub-surface seepage (ΣO). The mass balance relations in the above equations represent the governing components in the models we constructed with ISEE Systems™ *STELLA*® software.

Table 1
Mean annual climate statistics for selected sites in Petén, Guatemala.

Location	Lat (°N)	Long (°W)	Elevation (m)	Max T (°C)	Min T (°C)	Precip (mm yr ⁻¹)	Humidity (%)	Windspeed (m s ⁻¹)	Evap (mm mo ⁻¹)
El Paso ^a	17.21	90.19	115	29.5	20.7	1720	-	-	-
Flores ^b	16.93	89.88	123	31.4	20.0	1555	78	-	-
Flores ^c	-	-	-	31.9	21.6	-	77	1.7	-
Flores ^d	-	-	-	29.2	20.8	1510	-	-	90
Libertad ^b	16.78	90.12	125	31.3	19.7	1845	82	3.0	105
Poptun ^b	16.33	89.42	500	28.6	18.2	1850	82	-	-
Porvenir ^b	16.85	90.88	125	29.2	19.9	1860	-	-	-
San Andres ^b	16.97	89.92	150	32.8	19.7	1630	82	-	115

207	San Luis ^b	16.22	89.44	190	31.2	20.2	2125	81	-	-
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209 ^a Values for the period 1924-1934 (Lundell, 1937)

210 ^b Values for the period 1997-2002 (INSIVUMEH, unpublished data)

211 ^c Values for the period 1998-2003 compiled from daily data collected at the Flores/Santa Elena airport

212 ^d Values for the period 1973-1974 (Deevey et al., 1980)

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215 2.2. Hydrologic model equations

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217 The model calculates lake water balance through the volumetric addition of direct
 218 precipitation and runoff from the catchment, and subtraction of lake evaporation and
 219 outflow. The model assumes no temporal lag between surface and sub-surface flow to
 220 the lake and therefore, no distinction is made between surface and sub-surface
 221 components of catchment runoff (Vassiljev et al., 1998). In the absence of measurable
 222 surface drainage, lake water outflow is defined as a percentage of the overall lake volume
 223 lost through sub-surface seepage. The model calculates lake level (depth or elevation)
 224 and surface area from the derived volume and thus requires lake bathymetry and basin
 225 topography.

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227 For monthly time steps, the hydrologic model simulates direct precipitation over the lake
 228 from monthly precipitation inputs and the surface area of the lake derived from the area
 229 and depth curves. Evaporative loss from the lake is calculated with the empirical mass
 230 transfer formula of Brutsaert (1982):

$$231 \quad E = N u (e_{s-w} - e_{a-a}) \quad (3)$$

$$N = 3.367 \times 10^{-9} (A^{-0.05}) \quad (4)$$

where u is the wind speed (m s^{-1}) over the lake, e_{s-w} is the saturation vapor pressure at the surface water temperature, e_{a-a} is the actual vapor pressure at the overlying air temperature, and A is the surface area (m^2) of the lake. Saturation vapor pressure (millibars) is calculated from the equation of Murray (1967):

$$e_{s-w} \text{ or } e_{s-a} = 6.108 \text{ EXP} \left(\frac{17.27 T}{T + 237.7} \right) \quad (5)$$

where T is the temperature ($^{\circ}\text{C}$) of the lake water surface (e_{s-w}) or overlying air (e_{s-a}). Actual vapor pressure of the overlying air (e_{a-a}) follows from the saturation vapor pressure equation:

$$e_{a-a} = e_{s-a} \times h \quad (6)$$

where e_{s-a} is the saturation vapor pressure at the air temperature and h is the relative humidity expressed as a fraction.

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Potential catchment evapotranspiration (m month^{-1}) is derived from the empirical equations of Thornthwaite (1948):

$$PET = 0.016 \times f \times \left(\frac{10T}{H} \right)^a \quad (7)$$

$$H = \sum_{i=1}^{12} \left(\frac{T}{5} \right)^{1.5} \quad (8)$$

where f is an adjustment factor related to the number of daylight hours and latitude, T is the mean monthly air temperature between 0° and 26.5°C , H is the heat index, and a is a cubic function of the heat index:

$$a = 0.49 + 0.0179H - 0.000077H^2 + 0.000000675H^3 \quad (9)$$

Daylight hour and latitude adjustment factors are tabulated by Thornthwaite and Mather (1957). At temperatures above 26.5°C, potential evapotranspiration follows the function:

$$PET = -0.41585 + 0.03224T - 0.00043T^2 \quad (10)$$

where T is the mean monthly air temperature.

In the model, the rate of actual evapotranspiration (AET) is determined by potential evapotranspiration, the total area of the catchment covered with vegetation, and the availability of water stored within catchment soils. During wet months ($P > PET$) catchment evapotranspiration occurs at the potential rate and excess precipitation (P_e) recharges soil moisture stores:

$$P_e = \begin{cases} 0 & P \leq PET \\ P - PET & P > PET \end{cases} \quad (11)$$

Soil moisture (SM) increases to the defined maximum water capacity (MWC) of the soil as determined by soil texture and depth. If the amount of water within catchment soil stores exceeds the maximum water capacity, the excess is assumed to recharge the lake as surface and sub-surface inflow (ΣI):

$$\Sigma I = \begin{cases} 0 & SM + P_e \leq MWC \\ (P_e - MWC) + SM & SM + P_e > MWC \end{cases} \quad (12)$$

No excess precipitation is generated if catchment precipitation is equal to or less than potential evapotranspiration ($P \leq PET$). In such cases, soil moisture remains unchanged and overland flow and near-surface inflow are eliminated.

Evapotranspiration is reduced from the potential value if catchment precipitation falls short of vegetation water demand. Soil moisture loss (SM_{loss}) is thereby calculated as a function of precipitation and potential evapotranspiration:

$$SM_{\text{loss}} = \begin{cases} 0 & P \geq PET \\ |P - PET| & P < PET \end{cases} \quad (13)$$

It is important to note that any precipitation shortfall is assumed to be withdrawn from soil moisture stores at a rate proportional to the potential evapotranspiration until the soil moisture storage is exhausted.

2.3. Isotopic model equations

The isotopic mass-balance model calculates the isotopic composition of evaporative waters by the formula of Craig and Gordon (1965):

$$\delta_E = \frac{\alpha^* \delta_L - h_n \delta_A - \varepsilon_{\text{tot}}}{1 - h_n + 0.001 \varepsilon_{\text{kin}}} \quad (14)$$

where α^* is the inverse of the equilibrium isotopic fractionation factor ($1/\alpha$, $\alpha^* < 1$), h_n is the relative humidity normalized to the temperature of the lake surface water, δ_A is the isotopic composition of atmospheric moisture, ε_{kin} is the kinetic isotopic fractionation, ε_{tot} is the total (per mil) isotopic fractionation ($\varepsilon_{\text{eq}} + \varepsilon_{\text{kin}}$), and ε_{eq} is the equilibrium isotopic fractionation.

Normalized relative humidity is calculated from the saturation vapor pressure of the overlying air (e_{s-a}) and the saturation vapor pressure (e_{s-w}) at the surface water temperature:

$$h_n = h \times \frac{e_{s-a}}{e_{s-w}} \quad (15)$$

Atmospheric moisture (δ_A) is assumed to be at isotopic equilibrium with precipitation (Zimmerman et al., 1967, Zuber, 1983, Horita, 1990, and Gibson et al., 1993, 1999):

$$\delta_A = \delta_P - \varepsilon_{eq} \quad (16)$$

The isotopic mass-balance model calculates the equilibrium oxygen isotopic fractionation factor (α) and its inverse (α^*) from the equation of Majoube (1971):

$$\ln \alpha = 1137 T^{-2} - 0.4156 T^{-1} - 0.0020667 \quad (17)$$

where T is the temperature (Kelvin) of the lake surface water. The equilibrium isotopic fractionation of oxygen follows accordingly:

$$\varepsilon_{eq} = 1000(1 - \alpha^*) \quad (18)$$

Non-equilibrium fractionation of oxygen isotopes occurs during evaporation from the lake surface. This kinetic fractionation (ε_{kin}) is controlled by molecular diffusion and the moisture deficit ($1 - h$) over the lake surface (Merlivat and Jouzel 1979):

$$\varepsilon_{kin} = C \times (1 - h_n) \quad (19)$$

where C is the experimentally-derived isotopic fractionation value of 14.2‰ for oxygen (Vogt, 1976, and Araguás-Araguás et al., 2000) and h_n is the humidity normalized to the temperature of the lake surface water (Eq. 15).

The equation relating lake water $\delta^{18}\text{O}$ to the $\delta^{18}\text{O}$ of biogenic carbonate (ostracode shell calcite), is defined by Xia et al. (1997):

$$\delta_{\text{B}} = \delta_{\text{L}} - 0.179 T + 3.943 \quad (20)$$

where δ_{B} is the isotopic composition of the biogenic carbonate, δ_{L} is the $\delta^{18}\text{O}$ of the lake water (‰, SMOW), and T is the temperature of the lake water (°C).

2.4. Model input parameters

Initial model input parameters (Table 2) were derived from two decades of lake observation and historical climate data. Air temperature and precipitation inputs were derived from monthly average values recorded at four sites within ~80 km of Lake Salpetén (El Paso, Flores, Porvenir, and San Ignacio). Average monthly windspeed and relative humidity were calculated from daily values recorded at Flores, ~30 km from the lake. Discontinuous lakewater temperature measurements collected between 1980 and 1999 were used to calculate a proportional relationship between mean monthly air and surface water temperature.

Lake and catchment surface area were determined from topographic maps. Lake depth and volume were determined with the bathymetric estimates of Anselmetti et al. (2007). Catchment forest cover and soil water characteristics (maximum water capacity) were obtained from regional datasets (Zobler, 1986, Bouman et al., 1993, and Sader et al., 1994). Oxygen isotopic input values (δ_{L} , δ_{I} , and δ_{P}) were derived from measured values of lakewater, groundwater, and local precipitation, and interpolated values computed

using the Online Isotopes in Precipitation Calculator (*OIPC*) (<http://www.waterisotopes.org>) (Bowen and Wilkinson, 2002, and Bowen and Revenaugh, 2003). Because long-term data on $\delta^{18}\text{O}$ of regional precipitation are lacking, some error may be inherent within the interpolated δ_{P} values.

Table 2
Initial model input parameters for Lake Salpetén, Petén, Guatemala

Symbol	Description	Value	Data Source
V	Initial Lake Volume	$2.052 \times 10^7 \text{ m}^3$	Bathymetric Map Estimate
SM	Soil Moisture Stores	0	Assumed (End of Dry Season)
P	Precipitation ^a	$1755 \text{ mm year}^{-1}$	INSIVUMEH
O	Outflow (Percent of Total Volume)	$1.25\% \text{ month}^{-1}$	Deevey et al., 1980
δ_{L}	Isotopic Composition of Lake	4.1‰	Measured
δ_{P}	Isotopic Composition of Precipitation ^a	-4.0‰	Measured; OIPC Interpolations
δ_{I}	Isotopic Composition of Inflow ^a	-4.0‰	Measured; OIPC Interpolations
T_{Air}	Air Temperature ^b	25.3°C	INSIVUMEH
T_{Lake}	Lake Temperature ^b	28.5°C	Measured
h	Relative Humidity ^b	75%	INSIVUMEH
u	Wind Speed ^b	1.7 m s^{-1}	INSIVUMEH
MWC	Maximum Water Capacity ^c	0.03 m	Zobler, 1986; Bouman et al., 1993
-	Catchment Surface Area	$6.360 \times 10^6 \text{ m}^2$	Topographic Map Estimate
-	Percent of Catchment With Vegetation Cover	90%	Sader et al., 1994

^a Weighted mean annual value is given. The isotopic composition of inflow is assumed equal to precipitation.

^b Monthly parameter varied between and within model experiments. Mean annual value is given. (See Fig. 2 for monthly averages).

^c Assumes an average catchment soil profile depth of 0.6 m.

3. Results and discussion

All model simulations were performed with the *STELLA*® software using a fourth-order Runge-Kutta numerical approximation scheme. In the first series of model experiments, the modern climate and catchment parameter datasets were used to approximate the seasonal and long-term variability of Lake Salpetén water levels and lakewater $\delta^{18}\text{O}$. In the second sequence of model simulations, sediment core records of pollen-inferred forest cover change (Leyden, 1987 and Islebe, unpublished data) were incorporated into the model to investigate the impact of vegetation changes on basin hydrology over the last 4,000 years.

3.1. Model experiments utilizing modern catchment variables

In the first set of experiments, sub-surface outflow (seepage) was assigned a zero value and the lake was assumed to lose water only by evaporation. Such closed-basin simulations failed to approximate modern lake levels and exceeded the spillover height of the catchment determined from topographic maps (128 m a.s.l.). Sub-surface lake outflow was thereby progressively increased in a series of 360-month simulations until the modern level of Lake Salpetén was predicted accurately.

Modern lake levels were properly simulated only by including monthly sub-surface water losses (seepage) nearing 1.3% of the total lake volume, i.e. ~15% of the annual water budget. Deevey et al. (1980) identified comparable non-evaporative losses at Lakes Yaxhá and Sacnab (Fig. 1) during one year of monitoring in the early 1970s. The measured water losses, 120 cm at Lake Yaxhá and 135 cm at Lake Sacnab, represented

between 15 and 19% of the total volume of the lakes. In the absence of surface outflow, Deevey et al. (1980) suggested that this lost volume was “incipient groundwater” stored for some time within clay sediments of the lake floor.

In the sub-surface seepage configuration, the simulated lake reached hydrologic and isotopic steady-state conditions within ~15 model years (180 monthly time-steps). The simulated lake-level curve reproduced observed patterns of seasonal lake-level change (~0.9 m) with maximum stage between November and January and minimum stage between May and July (Fig. 2a). The model experiment also reproduced the mean summer month lakewater $\delta^{18}\text{O}$ (+4.1‰) and seasonal variations of $\pm 0.9\text{‰}$ (Fig. 2b).

Simulated evaporation using the Brutsaert formula (Eq. 3) averaged 110 mm month⁻¹. Although no direct evaporation measurement data exist for Lake Salpetén, modeled monthly evaporation is consistent with pan evaporation rates of 105 and 115 mm month⁻¹ reported at Libertad and San Andres, respectively (Table 1). Model-derived lake evaporation is slightly higher than the evaporation rates of 90 mm month⁻¹ measured at Lakes Yaxhá and Sacnab (Deevey et al. 1980). Modeled steady-state values of the $\delta^{18}\text{O}$ of evaporated water vary seasonally between -0.7‰ and -17.5‰ (Table 3).

Modern climate and catchment datasets were also modified to approximate the impacts of increased precipitation and changes in catchment vegetation cover on the hydrologic balance and lakewater $\delta^{18}\text{O}$ in Lake Salpetén. Simulations of approximately 2,000 model years were completed during each experiment. Between the 400th and 800th model year,

precipitation was increased or vegetation cover was decreased. Precipitation and vegetation cover were returned to the initial input values between the 1,200th and 1,600th model years. The length of each simulation was sufficient for the lake to establish isotopic and hydrologic steady state.

Table 3
Steady-state sample calculation of the isotopic composition of evaporated water

M^a	T_{Air}	T_{Lake}	h	$e_{\text{s-a}}$	$e_{\text{a-a}}$	$e_{\text{s-w}}$	h_n	α^{*b}	$\varepsilon_{\text{eq}}^c$	$\varepsilon_{\text{kin}}^c$	δ_L	δ_P	δ_A	δ_E
Eq. ^d	-	-	-	5	6	5	15	18	19	20	-	-	16	14
N	23.9	27.2	0.782	29.66	23.19	35.97	0.645	0.9909	9.12	5.04	3.42	-1.90	-11.02	-10.19
D	22.7	26.0	0.781	27.59	21.55	33.52	0.643	0.9908	9.21	5.07	3.40	-1.16	-10.37	-11.72
J	22.5	25.8	0.767	27.26	20.91	33.12	0.631	0.9908	9.23	5.24	3.44	-1.21	-10.44	-11.93
F	22.7	26.0	0.707	27.59	19.51	33.52	0.582	0.9908	9.21	5.94	3.50	-0.60	-9.81	-14.07
M	25.5	28.8	0.629	32.63	20.53	39.48	0.520	0.9910	8.99	6.82	3.63	1.56	-7.43	-17.14
A	27.1	30.4	0.619	35.86	22.20	43.29	0.513	0.9911	8.87	6.92	3.96	-2.84	-11.71	-11.85
M	28.1	31.4	0.671	38.02	25.51	45.83	0.557	0.9912	8.79	6.30	4.20	-3.24	-12.03	-9.39
J	27.2	30.5	0.751	36.07	27.09	43.54	0.622	0.9911	8.86	5.36	4.28	-5.17	-14.03	-3.26
J	26.0	29.3	0.795	33.61	26.72	40.64	0.658	0.9911	8.95	4.86	4.13	-3.46	-12.41	-4.49
A	26.3	29.6	0.800	34.22	27.37	41.35	0.662	0.9911	8.93	4.80	3.99	-3.29	-12.22	-4.90
S	26.2	29.5	0.802	34.01	27.28	41.11	0.664	0.9911	8.94	4.78	3.86	-5.88	-14.82	-0.16
O	25.4	28.7	0.800	32.44	25.95	39.25	0.661	0.9910	9.00	4.81	3.55	-2.88	-11.88	-7.09

^a The letter M denotes the month of the year. Model year simulations consist of a water year, beginning with the month of November.

^b $\alpha^* = 1/\alpha$ and $\ln \alpha = 24844T^{-2} - 76.248T^{-1} + 52.612$.

^c $\varepsilon_{\text{tot}} = \varepsilon_{\text{eq}} + \varepsilon_{\text{kin}}$.

^d Numbers correspond to the equation listed within the model description.

Figure 2: Rosenmeier et al. ↑

Fig. 2.

(a) Twelve-month model simulation of steady-state water level changes at Lake Salpetén (open circles), using historic data for inputs of rainfall, temperature, humidity and windspeed. Mean monthly lake level changes for other basins in Petén (Deevey et al., 1980) are plotted for comparison. No lake stage data are available for Salpetén. (b) Twelve-month steady-state simulation of lake water $\delta^{18}\text{O}$ (‰, SMOW) changes at Lake Salpetén (filled squares) versus lake level (open diamonds). The mean of lake water $\delta^{18}\text{O}$ values measured during the months of May, June, July and August is shown for comparison (open circle). Black bars indicate the full range of summer month $\delta^{18}\text{O}$ values.

Precipitation was first increased to values between 105 and 130% of the modern monthly dataset. Each successive 5% increase elevated the lake surface by nearly 1.2 m (Fig. 3a). Vegetation cover decreases of 15% produced lake level changes comparable in magnitude (1.3 m) to the 5% increases in precipitation (Fig. 3b). The principal effect of catchment vegetation loss is reduced evapotranspiration and increased surface and near-surface runoff to the lake (Eqs. 11 through 13). This effect is strongest in the later wet season months of August, September, and October, when soils are at maximum water capacity and any excess precipitation is converted directly to surface and near-surface runoff. Step-wise precipitation increases of 5% or vegetation losses of 15% and consequent average lake level changes of ~1.25 m produced a consistent 0.15‰ decrease in lakewater $\delta^{18}\text{O}$ (Figs. 3c and 3d). Combined increases in precipitation and reductions in vegetation cover increased the overall magnitude of simulated lake level rise and decreased lakewater $\delta^{18}\text{O}$ (Figs. 4a and 4b).

3.2. Uncertainty of model variables

We conducted additional experiments to determine the sensitivity of individual simulations to changes in relative humidity, windspeed and lake temperature. Relative humidity values 10% above modern and windspeed values 10% below modern produced lake surface elevation rises, through decreased evaporation, of ~0.9 m and respective lake water $\delta^{18}\text{O}$ decreases of 1.1 and 0.3‰ (Table 4). The effect of relative humidity on lakewater isotopic composition (δ_{L}) was amplified largely through calculated evaporated water $\delta^{18}\text{O}$ (δ_{E}) values (Eq. 14). Lake water temperatures 5% below modern, equivalent

Figure 3: Rosenmeier et al. ↑

Fig. 3.

Simulated response of lake level and lake water $\delta^{18}\text{O}$ to systematic changes in precipitation (a and b) and catchment vegetation cover (c and d). Heavy black lines delineate equilibrium lake level and $\delta^{18}\text{O}$, simulated using modern precipitation and modern catchment vegetation cover (90%).

Figure 4: Rosenmeier et al. ↑

501

502 Fig. 4.

503 Simulated lake level increase (a) and lake water $\delta^{18}\text{O}$ decrease (b) as a function of
 504 combined precipitation and catchment vegetation cover. Vegetation cover of 90%
 505 approximates the modern value.

506

to 1.5°C cooler, resulted in elevation of the lake surface by ~0.3 m and reduction of lakewater $\delta^{18}\text{O}$ by ~0.4‰, again through decreased evaporation (Table 4). Temperature fluctuations, however, are thought to have been negligible during the late Holocene (Hodell et al. 2012), and variations in lakewater $\delta^{18}\text{O}$ were thus probably dominated by changes in the relative rates (and isotopic composition) of hydrologic inputs and outputs.

Table 4
Sensitivity analysis of poorly constrained climate variables.

Parameter	Change	Model Result ^{a,b}
h	1% increase	δ_L decreases by 3.0%; lake level increases by 0.1%
h	10% increase	δ_L decreases by 30.0%; lake level increases by 1.0%
u	1% decrease	δ_L decreases by 0.8%; lake level increases by 0.1%
u	10% decrease	δ_L decreases by 9.0%; lake level increases by 0.8%
T_{water}	1% decrease	δ_L decreases by 2.0%; lake level increases by 0.3%
T_{water}	5% decrease	δ_L decreases by 10.0%; lake level increases by 0.3%
δ_p	1% decrease	δ_L decreases by 1.0%; lake level remains unchanged
δ_p	10% decrease	δ_L decreases by 8.0%; lake level remains unchanged

^a Annual average at steady-state (equilibrium condition) is listed.

^b Any 10% decrease in u is equivalent to ~0.35‰ (SMOW). Any 1.0% increase in lake level is equivalent to ~1.0 m.

Modeled lakewater $\delta^{18}\text{O}$ is also sensitive to changes in the isotopic composition of precipitation (Table 4). This was expected and has been demonstrated in studies of modern oxygen isotope values in rainfall and surface waters across Central America (Lachniet and Patterson, 2009). As previously noted, no long-term data exist for the $\delta^{18}\text{O}$ of regional precipitation, which may contribute error to the interpolated δ_p values. Trace element (Mg and Sr) concentrations in sediment cores from Lake Salpetén, however,

track $\delta^{18}\text{O}$ fluctuations, suggesting that $\delta^{18}\text{O}$ values reflect watershed hydrology rather than changes in the isotopic composition of rainfall (Rosenmeier et al., 2002b).

3.3 Model experiments utilizing paleoenvironmental variables

Model experiments utilizing modern climate and catchment parameters demonstrate that both precipitation amount and percent vegetation cover influence lake level and lakewater $\delta^{18}\text{O}$. Palynologically documented changes in forest cover were therefore incorporated within a series of 4,000-year model simulations to evaluate the effect of late Holocene vegetation changes on the hydrology of Lake Salpetén. Previous work on sediment cores from Lake Salpetén showed that biogenic carbonate $\delta^{18}\text{O}$ values declined between 3,300 and 2,400 cal yr BP (Fig. 5a) (Rosenmeier et al., 2002a), coinciding with palynologically documented forest loss (Fig. 5b) (Leyden, 1987). Minimum $\delta^{18}\text{O}$ values, suggesting high water levels, occurred between 2,400 and 1,800 cal yr BP. Independent evidence for high water level is provided by radiocarbon dates on aquatic gastropod shells from exposed lake sediments that lie between 1.0 and 7.5 m above the present lake surface (Rosenmeier et al., 2002b). Oldest dates indicate the lake filled to a depth >7.0 m above modern lake stage by ~2,800 cal yr BP, but most dates indicate episodes of high lake level between 2,000 and 1,200 cal yr BP (Fig. 5a). These radiocarbon ages coincide with reduced sediment $\delta^{18}\text{O}$ values and support the inference for increased water input to Lake Salpetén. Forest loss may have decreased evapotranspiration and soil moisture storage in the watershed, thereby increasing catchment water yield and transport of surface and ground waters with lower $\delta^{18}\text{O}$ values to the lake. Increased delivery of input waters

would have increased lake volume, decreased the proportion of the hydrologic budget lost to evaporation, and decreased lakewater $\delta^{18}\text{O}$. Alternatively, high water levels and $\delta^{18}\text{O}$ minima may reflect increased precipitation and reduced evaporation beginning ~3,300 years ago.

Model-derived late Holocene changes in lake level and lakewater $\delta^{18}\text{O}$, driven by shifts in catchment vegetation cover, are consistent with paleohydrologic conditions inferred from sediment core biogenic carbonate $\delta^{18}\text{O}$ (Figs. 6a and 6b). Simulations integrating pollen-derived vegetation cover explain 75% of the variance in the sediment core biogenic carbonate $\delta^{18}\text{O}$ (Figs. 6c and 6d). Nonetheless, these simulations do not encompass the full magnitude of measured sediment core $\delta^{18}\text{O}$ values and fail to account for abrupt $\delta^{18}\text{O}$ increases centered at 1,500, 1,100, 900, 500, and 200 cal yr BP.

Figure 5: Rosenmeier et al. ↑

574

575 Fig. 5.

576 (a) Oxygen isotopic composition of ostracod valves (open circles) in Lake Salpetén
 577 sediment cores (Rosenmeier et al., 2002a) and lake level high stands inferred from sub-
 578 aerial lacustrine deposits ~1.0 to 7.5 m above modern stage (Rosenmeier et al., 2002b)
 579 versus calendar age. Oxygen isotopic data were smoothed with a 5-point running mean.
 580 Age of the lake level high stands was determined by radiocarbon dating of aquatic
 581 gastropods, corrected for hard-water-lake error. (b) Relative abundance of high forest
 582 pollen taxa from Lake Salpetén sediment cores (Leyden, 1987; Islebe, unpublished data).
 583 High forest taxa include species of the families Moraceae and Urticaceae.

584

585

Figure 6: Rosenmeier et al. ↑

Fig. 6.

(a) Pollen-based simulation of catchment vegetation cover changes and (b) consequent lake level (black line) and lake water $\delta^{18}\text{O}$ (gray line) variations. The thickness of the lines results from the seasonal variations in climatic inputs. (c) Comparison of the Lake Salpetén sediment core biogenic carbonate $\delta^{18}\text{O}$ record (open circles) and simulated biogenic carbonate $\delta^{18}\text{O}$ (gray line). Model ages and sediment core ages were converted to calendar years before present. (d) Linear regression of model and sediment core $\delta^{18}\text{O}$ values.

Minimum sediment core $\delta^{18}\text{O}$ values in Lake Salpetén coincide with deposition of a thick erosional clay unit that has been attributed to human-induced deforestation (Anselmetti et al., 2007). Soil erosion and decreased storage of soil moisture may have accelerated inflow to the lake of water with lower $\delta^{18}\text{O}$ values. Accordingly, additional model experiments simulated reduced soil thickness, from soil erosion, and decreased water storage capacity. These changes accounted for <25% of the difference between the sediment record and model simulation of the $\delta^{18}\text{O}$ minimum. Even if changes in the soil moisture budget could account for the offset between the model and core data, decreased water storage capacity of the catchment cannot explain the periodic reductions of inflow inferred from abrupt sediment core $\delta^{18}\text{O}$ increases. Either additional variability in vegetation cover or precipitation change is therefore required to explain these abrupt changes in sediment core $\delta^{18}\text{O}$.

It is conceivable that undetected rapid vegetation changes may have altered lake hydrology, and thereby lake water isotopic composition, as pollen from the Lake Salpetén sediment core was sampled at relatively broad intervals. Forest cover reconstructions thus lack the resolution to distinguish the short-term fluctuations discernible in the high-temporal-resolution $\delta^{18}\text{O}$ profile. The full variability of the sediment core $\delta^{18}\text{O}$ record was recreated using synthetic vegetation change values (Fig. 7a). Model simulations that accounted for the full variability in the $\delta^{18}\text{O}$ record required reduction of forest cover to <5% of the total watershed area and afforestation exceeding 100% coverage of the basin (Fig. 7b). Some of the variability incorporated in this artificial vegetation record is clearly implausible. It is therefore likely that some of the change observed in the Lake

Salpetén $\delta^{18}\text{O}$ record reflects the influence of climate variations, particularly shifts in precipitation amount.

Model simulations did a better job of reproducing the $\delta^{18}\text{O}$ values from the sediment record when they incorporated changes in both vegetation cover and precipitation. For example, minimum $\delta^{18}\text{O}$ values between 2,400 and 1,800 cal yr BP were successfully reproduced by the model when using the sediment core, pollen-based estimates of vegetation cover and superimposition of an average increase in precipitation of about 15% (Figs. 7c and 7d). Simulation of the increases in sediment core $\delta^{18}\text{O}$ centered at ~3,300, 2,900, 500, and 200 cal yr BP necessitated precipitation decreases of at least 10% relative to modern. Simulation of sediment core $\delta^{18}\text{O}$ values between ~1,500 and 800 cal yr BP required inclusion of even drier conditions.

A comparison of the $\delta^{18}\text{O}$ record of ostracods from Lake Salpetén and the Yok Balum speleothem from Belize (Fig. 8) is useful to distinguish how much of the Salpetén $\delta^{18}\text{O}$ signal is climate related. Both records are from the southern Lowlands and the speleothem $\delta^{18}\text{O}$ record is interpreted to reflect changes in the $\delta^{18}\text{O}$ of precipitation, enhanced by evaporation in the cave (Kennett et al., 2012). The Yok Balum record only extends to 2,500 yrs BP, but the lowest $\delta^{18}\text{O}$ values in Lake Salpetén during the Late Preclassic are not reflected by similarly low values in Yok Balum $\delta^{18}\text{O}$. This supports our model results that the Lake Salpetén record was influenced by increased runoff from deforestation during the Preclassic Period. The $\delta^{18}\text{O}$ trends in both records, however, are broadly similar during the Classic and Postclassic Periods, suggesting that the Salpetén

record also reflects changes in climate, which appear coincident with Maya cultural transitions (Kennett et al., 2012).

Inferred late Holocene precipitation decreases at Lake Salpetén also agree with records of drying from the northern Yucatán Peninsula and suggest that shifts in moisture availability were regional in extent. Model-derived precipitation at Lake Salpetén was relatively high during the Middle and Late Preclassic periods (Fig. 9a). The boundary between the Preclassic and Classic was marked by an abrupt decline in precipitation that persisted throughout the Classic. The sediment record from Lake Punta Laguna (Curtis et al., 1996, and Hodell et al., 2007) displayed a similar abrupt increase in $\delta^{18}\text{O}$ after ~1,800 cal yr BP that remained high throughout the Classic period, suggesting protracted dry conditions (Fig. 9b). In Lake Chichancanab, $\delta^{18}\text{O}$ increased abruptly at ~1,200 cal yr BP and marked the onset of a nearly 200-year period of droughts in the Terminal Classic and earliest Postclassic periods (Fig. 9c) (Hodell et al., 1995, 2001, 2005).

Figure 7: Rosenmeier et al. ↑

Fig. 7.

(a) Synthetic catchment vegetation cover change record (black bars) used to simulate the variability observed in sediment core $\delta^{18}\text{O}$ values from Lake Salpetén. Pollen-based reconstruction of vegetation change (open circles) is shown for comparison. (b) Comparison of the Lake Salpetén sediment core biogenic carbonate $\delta^{18}\text{O}$ record (open circles) and biogenic carbonate $\delta^{18}\text{O}$ values (gray line) resulting from use of the synthetic vegetation record. (c) Precipitation changes (percent increase or decrease from modern) used to further alter the hydrologic and isotopic mass balance model of Lake Salpetén. (d) Comparison of the sediment core biogenic carbonate $\delta^{18}\text{O}$ record and biogenic carbonate $\delta^{18}\text{O}$ values resulting from the superimposition of precipitation changes and pollen-based simulations of catchment vegetation change.

Rosenmeier et al. Figure 8

Fig. 8

Comparison of the $\delta^{18}\text{O}$ records of ostracods from Lake Salpeten cores SP 1 (black line) and SP2 (blue line) and the Yok Balum speleothem from Belize (red line) (Kennett et al., 2012). Note that time runs from left (older) to right (younger).

681

682

Figure 9: Rosenmeier et al.

683

684

685 Fig. 9.

686 (a) Regional precipitation changes inferred from the hydrologic and isotopic mass balance
 687 model of Lake Salpetén and oxygen isotope records from (b) Punta Laguna (Curtis et al.,
 688 1996) and (c) Chichancanab (Hodell et al, 1995) versus age and Maya cultural periods
 689 (Rice and Rice, 1990): Hist. = Historic, Post. = Postclassic, T = Terminal Classic, LPRC
 690 = Late Preclassic, MPRC = Middle Preclassic.

691

4. Conclusions

Model simulations of hydrology and oxygen isotopic composition in Lake Salpetén, Guatemala have broad implications with respect to isotope-based paleolimnological inferences for past climate in the Maya Lowlands. Several decades ago it became apparent that pollen could not be utilized to infer late Holocene climate change on the Yucatán Peninsula, given that the ancient Maya dramatically transformed regional forests through their agricultural and urbanization activities. Alternative paleoclimate proxies were sought and stable oxygen isotope measures on the shells of fossil aquatic organisms appeared to offer the possibility for high-resolution reconstruction of past relations between evaporation and precipitation, free from human influence. Model simulations of hydrology and $\delta^{18}\text{O}$ in Lake Salpetén, however, indicate that 75% of the measured sediment core biogenic carbonate $\delta^{18}\text{O}$ variability is explained by palynologically documented changes in past catchment forest cover. Ancient Maya deforestation, which precludes climate inference using pollen, similarly complicates traditional interpretation of shell $\delta^{18}\text{O}$ data from lake cores in the area. Nevertheless, the regional climate signal is preserved in the Lake Salpetén lake sediments, as vegetation changes alone failed to reproduce the full range of variability observed in the sediment core $\delta^{18}\text{O}$ record. Simulations reproduced sediment core $\delta^{18}\text{O}$ values accurately only when changes in both forest cover and rainfall were incorporated into the model.

It is important to note that the relative contributions of vegetation cover and rainfall amount, as determinants of sediment core carbonate $\delta^{18}\text{O}$ values, probably varied through

time and space in lakes of the Maya Lowlands. The Yucatán Peninsula hosts a range of vegetation types, from low-stature, deciduous forest in the north, to high-stature evergreen vegetation in the south (Flores and Carvajal, 1994). The hydrologic consequence of felling these different vegetation types, which occupy habitats characterized by different rainfall amounts, probably varies considerably. Furthermore, it is likely that there was substantial spatial and temporal variability in human-mediated deforestation throughout the region. There exists a pronounced gradient in rainfall across the peninsula today and there is emerging evidence that past shifts in precipitation amount were not the same among sites in the area. For example, Douglas et al. (2015) reported that drought intensity during the Terminal Classic Period (AD ~800 to 950) was more intense in the southern Maya Lowlands, relative to the northern Maya Lowlands. Finally, local topography, along with factors such as the ratio between watershed drainage area and lake surface area, influence how landscape clearance alters local hydrology. The steep, haystack-hill topography of Petén, Guatemala, promotes rapid runoff when the landscape is denuded, whereas the relatively flat topography at sites farther north on the peninsula might experience more downward seepage of water following vegetation removal. This suggests that fluctuations in shell $\delta^{18}\text{O}$ values from lakes in the drier, flatter northern lowlands might have been driven more by shifts in past rainfall amount than human land use. Such speculation remains to be tested at other sites, using the modeling approach.

738

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740

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